(2007) believe that circulation patterns such as the Beaufort Gyre, which in the past helped to maintain old ice in the Arctic Basin, are now acting to export ice, as the multi-year ice is no longer surviving the transport through the Chukchi and East Siberian Seas.

According to DeWeaver (2007): ''Recognizing the need to incorporate AO variability into considerations of recent sea ice decline, Lindsay and Zhang (2005) used an ocean-sea ice model to reconstruct the sea ice behavior of the satellite era and identify separate contributions from ice motion and thermodynamics. Similar experiments with similar results were also reported by Rothrock and Zhang (2005) and Koberle and Gerdes (2003).'' Rothrock and Zhang (2005, cited in Serreze et al. 2007, pp. 1,533–1,536), using a coupled ice-ocean model, argued that although wind forcing was the dominant driver of declining ice thickness and volume from the late 1980s through the mid-1990s, the ice response to generally rising air temperatures was more steadily downward over the study period (1948 to 1999). ''In other words, without wind forcing, there would still have been a downward trend in ice extent, albeit smaller than that observed'' (Serreze et al. 2007, pp. 1,533–1,536). Lindsay and Zhang (2005, cited in Serreze et al. 2007, pp. 1,533–1,536) came to similar conclusions in their modeling study: ''Rising air temperature reduced ice thickness, but changes in circulation also flushed some of the thicker ice out of the Arctic, leading to more open water in summer and stronger absorption of solar radiation in the upper (shallower depths of the) ocean. With more heat in the ocean, thinner ice grows in autumn and winter.''

Changes in Oceanic Circulation

According to Serreze et al. (2007, pp. 1,533–1,536), it appears that changes in ocean heat transport have played a role in declining Arctic sea ice extent in recent years. Warm Atlantic waters enter the Arctic Ocean through the Fram Strait and Barents Sea (Serreze et al. 2007, pp. 1,533–1,536). This water is denser than colder, fresher (less dense) Arctic surface waters, and sinks (subducts) to form an intermediate layer between depths of 100 and 800 m (328 and 2,624 ft) (Quadfasel et al. 1991) with a core temperature significantly above freezing (DeWeaver 2007; Serreze et al. 2007, pp. 1,533–1,536). Hydrographic data show increased import of Atlantic-derived waters in the early to mid-1990s and warming of this inflow (Dickson et al. 2000; Visbeck et al. 2002). This trend has continued, characterized by pronounced pulses of warm inflow (Serreze et al. 2007, pp. 1,533–1,536). For example, strong ocean warming in the Eurasian Basin of the Arctic Ocean in 2004 can be traced to a pulse entering the Norwegian Sea in 1997–1998 and passing through Fram Strait in 1999 (Polyakov et al. 2007). The anomaly found in 2004 was tracked through the Arctic system and took about 1.5 years to travel from the Norwegian Sea to the Fram Strait region, and an additional 4.5–5 years to reach the Laptev Sea slope (Polyakov et al. 2007).

Polyakov et al. (2007) reported that mooring-based records and oceanographic surveys suggest that a new pulse of anomalously warm water entered the Arctic Ocean in 2004. Further Polyakov et al. (2007) stated that: ''combined with data from the previous warm anomaly * * * this information provides evidence that the Nansen Basin of the Arctic Ocean entered a new warm state. These two warm anomalies are progressing towards the Arctic Ocean interior * * * but still have not reached the North Pole observational site. Thus, observations suggest that the new anomalies will soon enter the central Arctic Ocean, leading to further warming of the polar basin. More recent data, from summer 2005, showed another warm anomaly set to enter the Arctic Ocean through the Fram Strait (Walczowski and Piechura 2006). These inflows may promote ice melt and discourage ice growth along the Atlantic ice margin (Serreze et al. 2007, pp. 1,533–1,536).

Once Atlantic water enters the Arctic Ocean, the cold halocline layer (CHL) separating the Atlantic and surface waters largely insulates the ice from the heat of the Atlantic layer. Observations suggest a retreat of the CHL in the Eurasian basin in the 1990s (Steele and Boyd 1998, cited in Serreze et al. 2007, pp. 1,533–1,536). This likely increased Atlantic layer heat loss and ice-ocean heat exchange (Serreze et al. 2007, pp. 1,533–1,536), which would serve to erode the edge of the sea ice on a year-round basis (C. Bitz, in litt. to the Service, November 2007). Partial recovery of the CHL has been observed since 1998 (Boyd et al. 2002, cited in Serreze et al. 2007, pp. 1,533–1,536), and future behavior of the CHL is an uncertainty in projections of future sea ice loss (Serreze et al. 2007, pp. 1,533–1,536).

Synthesis

From the previous discussion, surface air temperature warming, changes in atmospheric circulation, and changes in oceanic circulation have all played a role in observed declines of Arctic sea ice extent in recent years.

According to DeWeaver (2007): ''Lindsay and Zhang (2005) propose a three-part explanation of sea ice decline,'' which incorporates both natural AO variability and warming climate. In their explanation, a warming climate preconditions the ice for decline as warmer winters thin the ice, but the loss of ice extent is triggered by natural variability such as flushing by the AO. Sea ice loss continues after the flushing because of the sea-ice albedo feedback mechanism which warms the sea even further. In recent years, flushing of sea ice has continued through other mechanisms despite a relaxation of the AO since the late 1990s. The sea-ice albedo feedback effect is the result of a reduction in the extent of brighter, more reflective sea ice or snow, which reflects solar energy back into the atmosphere, and a corresponding increase in the extent of darker, more absorbing water or land that absorbs more of the sun's energy. This greater absorption of energy causes faster melting, which in turn causes more warming, and thus creates a self-reinforcing cycle or feedback loop that becomes amplified and accelerates with time. Lindsay and Zhang (2005, p. 4,892) suggest that the sea-ice albedo feedback mechanism caused a tipping point in Arctic sea ice thinning in the late 1980s, sustaining a continual decline in sea ice cover that cannot easily be reversed. DeWeaver (2007) believes that the work of Lindsay and Zhang (2005) suggests that the observed record of sea ice decline is best interpreted as a combination of internal variability and external forcing (via GHGs), and raises the possibility that the two factors may act in concert rather than as independent agents.

Evidence that warming resulting from GHG forcing has contributed to sea ice declines comes largely from model simulations of the late 20th century climate. Serreze et al. (2007, pp. 1,533–1,536) summarized results from Holland et al. (2006, pp. 1–5) and Stroeve et al. (2007, pp. 1–5), and concluded that the qualitative agreement between model results and actual observations of sea ice declines over the PM satellite era is strong evidence that there is a forced component to the decline. This is because each of these models would be in its own phase of natural variability and thus could show an increase or decrease in sea ice, but the fact that they all show a decrease indicates that more than natural variability is involved, i.e., that external forcing by GHGs is a factor. In addition, the model results do not show a decline if they are not forced with the observed GHGs. Serreze et al.